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Shear Velocity Structure in the Eastern United States

from the Inversion of Surface Wave

Group and Phase Velocities

by

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Abstract

Single-station group velocities at periods between 4 and 70 seconds and single-station phase velocities at periods between 10 and 50 seconds are obtained for a broad region of the eastern United States for both Rayleigh and Love waves. These data were obtained by the single-station method using earthquakes with known fault-plane solutions in the central United States. The group velocity data are combined with earlier acquired attenuation data to yield Q values for Rayleigh and Love waves in the eastern United States. Inversions of the dispersion curves for shear-velocity models of the crust and upper mantle indicate that all of the data can be explained satisfactorily by a single model which is isotropic in its elastic properties. This model includes a surface layer of low-velocity sediments, and is characterized by upper crustal shear-wave velocities which increase from 3.4 to 3.7 km/sec with increasing depth, by lower crustal shear-wave velocities between 4.0 and 3.8 km/sec, and by an upper-most mantle shear-wave velocity of about 4.8 km/sec. The fundamental-mode data of this study do not require a low-velocity zone in either the crust or upper mantle. Higher-mode velocities and wave forms, although having potentially greater resolving power than the fundamental-mode data, were too inconsistent to contribute to the elucidation of detailed features of the crust-upper mantle model.

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Introduction

Surface wave group and phase velocities have provided important information on the properties of the crust and upper mantle in various regions of the Earth. Their usefulness is partly due to their sensitivity to shear wave structure; thus they can provide information which is often difficult to obtain from body wave studies. A second advantage is that no special assumptions or procedures are necessary to invert surface wave data for models which include low-velocity zones. Most surface wave studies of crust-upper mantle structure have employed either phase or group velocities. Studies by Bloch et el. (1969) and Yu and Mitchell (1978) have, however, emphasized the desireability of employing both phase and group velocities to invert for structure.

Surface waves are most useful for studying structure in regions that are devoid of large-scale lateral property changes. Results of seismic refraction and other geophysical studies suggest that the eastern and central portions of North America are such suitable regions. Crust-upper mantle models for those areas have previously been obtained by Brune and Dorman (1963) using Love and Rayleigh waves crossing the Canadian shield, by McEvilly (1964) also using Love and Rayleigh waves in the central United States, and by Mitchell (1965) who used only Rayleigh waves to study a broader region of the east-central United States. Biswas and Knopoff (1974) used long-period Rayleigh waves to obtain upper mantle models for the central United States, as well as for other regions of North America. In later sections, we will refer particularly to the models of McEvilly (1964) since they were obtained for a region which corresponds most closely with the

region of the present study. All of the above references employed only phase velocities. We have used group velocity, as well as phase velocity, data in an attempt to obtain a new, more detailed, crust-upper mantle model for the east-central United States.

An interesting feature of the study by McEvilly (1964), was that a single isotropic model for the crust and upper mantle could not be found which would explain both the Love wave and Rayleigh wave data. He gave three possible explanations for that result: (1) that the upper mantle or/and lower crust are anisotropic in their elastic properties, (2) that systematic errors occur in the fundamental-mode phase velocities due to interference by higher modes, and (3) that his model, which included three crustal layers and four upper mantle layers was too simplified to completely explain his data. Thatcher and Brune (1969) suggested that the second of these explanations was likely to be the correct one. Boore (1969), however, argued that higher-mode interference would produce a large scatter, but no uniform bias, if data were obtained from an ensemble of sources.

Two developments since the study of McEvilly (1964) have led us to attempt to obtain a new shear-velocity model for the central and eastern United States. First, a new data set including both group and phase velocities has been obtained using central United States earthquakes with known fault-plane solutions. Second, inversion techniques are now available which permit us to consider models which are more complicated than those of McEvilly (1964)

We will invert the fundamental-mode Love- and Rayleigh-wave data, both simultaneously and separately, to investigate the possibility of anisotropy in crust-upper mantle properties. This approach of using isotropic inversion methods to study anisotropy is not strictly valid (Crampin, 1970, 1976). It will, however, provide an indication of the severity of any anisotropy

that might be present. In addition, synthetic seismograms, incorporating
the fundamental mode and several higher modes will also be constructed in
an attempt to investigate features of the resulting models which cannot
be resolved using only the fundamental mode.

Sources

A fault-plane solution for only one central United States earthquake,
that of November 9, 1968, has been determined using teleseismic P-waves
(Stauder and Nuttli, 1970). Although numerous other earthquakes have
occurred there, especially in the New Madrid seismic zone, they have been
too small to be amenable to study by the classical method of teleseismic
P-wave first motions. Mitchell (1973a,b) however, showed that surface
waves are a valuable tool for determining depths and fault-plane solutions
for relatively small earthquakes in that region in studies of the southeastern
Missouri earthquake of October 21, 1965. He adopted a method similar to
that devised by Tsai and Aki (1969). The use of surface waves has made
it possible to obtain depths and fault-plane solutions for several central
and eastern United States earthquakes (Herrmann, 1974, 1978). A number
of these events will be used as sources of the surface data in the present
study.

Data

Our data were obtained for paths between earthquake sources in the central United States (Table 1 and Figure 1) and numerous stations in the central and eastern United States. Phase velocities were determined using the single-station method (Brune et al., 1960). By using earthquakes with known fault-plane solutions, the phase and group velocities can be corrected for the initial phase and group delay at the source (Knopoff and Schwab, 1968).

The group velocities of this study were obtained using the multiple-filter method (Dziewonski et al., 1969; Herrmann, 1973). Spectral amplitudes associated with the fundamental mode can be separated from higher modes and noise; thus a fundamental-mode group velocity value can be obtained which is not contaminated by higher-mode interference.

The group velocities of this study were obtained as by-products of source mechanism and surface wave attenuation determinations (Herrmann, 1974). Those determinations included data from numerous paths to seismograph stations in both eastern and western North America. That body of data, after correction for the group delay at the source, is plotted in Figure 2. The large scatter in the data can be attributed to several factors, including regional variations in surface wave velocities, the inclusion of data with low signal-noise ratios, possible mode misidentification, and incorrect adjustment for initial phase due to departures of the actual focal mechanism from the assumed fault-plane solution.

In order to obtain a reliable data set pertaining to the eastern United States, the data of Figure 2 have been windowed in several ways. An azimuth window (340° - 135°) and a distance window (300 - 2000 km) were

first applied to the data. The azimuth window restricts our region of study to a section of the United States centered south of the Great Lakes, and omits data for paths in the western and southern United States where significant changes in shear wave structure may occur (Biswas and Knopoff, 1974). The distance window removes data obtained at short distances where possible incorrect initial phase assumptions cause the largest errors in phase and group velocity determinations, and at large distances where effects of lateral structural changes, such as surface wave lateral refraction and multi-pathing are most severe. The data have also been windowed according to their amplitudes. For each event, data having spectral amplitudes smaller than one-fifth the maximum amplitude at the given period were excluded. This process removes data which are associated with nodes in the surface wave source radiation patterns, and which consequently have a low signal-noise ratio. In addition, a few group velocity points which are clearly removed from the main trend of the data, and probably correspond to noise or higher-mode maxima rather than to the fundamental mode, were deleted from the final data set. The windowed data appear in Figure 3. It is clear that a considerable reduction of scatter has resulted from the windowing process.

The great preponderance of surface wave paths are restricted to a region which extends northward from the source region to the Great Lakes and eastward from the source region to the Appalachian Mountains (see Figure 1). A few short segments of path extend into Canada, to the east coast, and to the southeastern United States. Therefore our data pertain largely to the northeastern and north central United States. It is worthwhile to inquire about the extent of lateral variations in elastic properties through these regions since significant lateral variations

could bias our results. We have addressed this question by considering pertinent results of earlier surface wave studies, seismic refraction studies, and investigations of Pn travel-times.

The most useful surface wave results bearing on this question are those of Pilant (1972). Those data appear as contour maps of Rayleigh wave phase velocities across the United States. Across the region of the present study Pilant's velocities vary between 3.40 and 3.55 km/sec at a period of 16 seconds, between 3.55 and 3.70 km/sec at a period of 20 seconds, between 3.70 and 3.80 km/sec at a period of 25 seconds, between 3.90 and 4.00 km/sec at a period of 33 seconds, between 3.95 and 4.05 km/sec at a period of 41 seconds, between 4.00 and 4.10 km/sec at a period of 51 seconds, and between 4.05 and 4.10 km/sec at a period of 64 seconds. Those variations are comparable to the uncertainty of our observations. In general, Pilant's maps suggest that although some lateral variations occur across the region of interest, they are relatively minor.

Seismic refraction studies in the central United States (e.g. Stewart, 1968, and Ocala and Meyer, 1973) as well as more easterly portions of the central plains (Steinhart and Meyer, 1961) indicate crustal thicknesses between 40 and 45 km. In the New York-Pennsylvania region, however, a crustal thickness of 36 km was obtained (Katz, 1955). Along the eastern seaboard, and outside of our region of interest, the crust apparently thins considerably. Pn velocities, as compiled by Herrin and Taggart (1965), are relatively uniform throughout our region of study, varying only between 8.1 and 8.2 km/sec.

Phase velocities were determined using the better recorded events of Table 1. Spectral phases were obtained, and corrected for the initial phase at the source. Phase velocities were calculated using the single-station method (Brune et al., 1960). In using that method, a family of phase velocity

curves are determined, each corresponding to a different integer N, which occurs because there is an uncertainty of 2Nπ in the observed phases. The value of the integer has been obtained by comparing the resulting phase velocities for various N values to the phase velocities of McEvilly (1964) and Biswas and Knopoff (1974) at longer periods.

Thatcher and Brune (1969) have pointed out that the velocities of long-period, fundamental-mode Love waves can be in error if there is significant interference with higher-mode Love waves. Difficulties due to interference with higher modes are apt to be most severe if the distance between the source and receiver includes a segment of oceanic path. Although there are no portions of oceanic path in the present study, it is still worthwhile to investigate the level of excitation of higher modes relative to that of the fundamental mode. Spectral amplitudes observed at a recording station will depend upon the depth and focal mechanism of the earthquake source, as well as upon the elastic and anelastic properties of the crust and upper mantle between the source and receiver. Spectral amplitudes for the fundamental Love mode and nine higher modes were computed using the formulation of Levshin and Yanson (1971) for the case of the earthquake of October 21, 1965 (Figure 4). The crust-upper mantle model used in the calculations is that obtained from Rayleigh waves as described in the following section. Spectra were determined for two source depths: 4 km, as obtained for this earthquake by Mitchell (1973a), and 15 km. The latter depth was chosen in order to illustrate the effect which variations in source depth have on the spectra. In both cases, at periods greater than about 10 seconds, the higher-mode amplitudes are down by more than an order of magnitude below those of the fundamental mode. We expect therefore, that our Love

wave data will not be significantly biased by the presence of higher modes.

Uncertainties in surface wave group and phase velocities can be determined from travel-time uncertainties associated with digitizing errors, inexact estimates of the initial phase at the source, uncertainties in origin time and source location, the effects of source finiteness, and the limited precision with which we can read group and phase travel times. We estimate the maximum uncertainties in the group and phase travel times to be about 8 seconds. At a distance of 1000 km, this time uncertainty leads to velocity uncertainties between 0.1 and 0.15 km/sec. The data points of Figure 3 exhibit a spread of about that much. The data used for inversion and their standard deviations are plotted in Figure 5. The standard deviations were calculated for all periods at which there were three or more data points. In cases where two data points were available, a standard deviation of 0.1 was used, and in cases where only one data point was obtained, a standard deviation of 0.15 was used.

Table 2 presents the mean group and phase velocities obtained in this study. Combining the group velocity data with values for surface wave attenuation coefficients (Mitchell, 1973a; Herrman and Mitchell, 1975) for the eastern United States allows us to tabulate Q values for Rayleigh and Love waves in that region. These values are also listed in Table 2.

Our observed group and phase velocity data are compared with theoretical values for three models of McEvilly (1964) in Figure 5. That set of models was derived from phase velocity data only and includes (1) a model which is isotropic at all depths, (2) a model which is anisotropic only in the upper mantle, and (3) a model which is anisotropic in both the upper mantle and lower crust. The last of

these models provided the best fit to McEvilly's data. The comparisons
in Figure 5 indicate that the phase velocities for McEvilly's model
which is anisotropic in both the lower crust and upper mantle fit the
present set of phase velocity data quite well. This result indicates
that the phase velocity data of McEvilly (1964) were not severely
biased due to higher-mode interference.

The Rayleigh wave group velocities predicted by the same model also
fit our observed data well, except at the shortest periods. The
lower observed group velocities at those periods suggest that a
low-velocity sedimentary layer is required for any eastern United
States model. None of the Love wave group velocity curves predicted
by McEvilly's models correspond very well to our observed group
velocities. Fitting these data, as well as the phase velocities
and Rayleigh wave group velocities, will apparently require a more
complicated model for the eastern United States than any which is
presently available.

Models Obtained From Inversion

Shear wave models were obtained from the surface wave data using modern inversion theory (Bachus and Gilbert, 1970) in stochastic form (Der et al., 1970; Jordan and Franklin, 1971). The program we have used for the inversions was written by W. L. Rodi and employs the surface wave algorithm of Harkrider (1964) and efficient methods for computing partial derivatives (Rodi et al., 1975). We have assumed that the compressional velocities and densities are known and have solved only for the shear velocities. Several compressional wave crustal models for the eastern United States are available. Two of these (McCamy and Meyer, 1966; Mitchell and Hashim, 1977) pertain to the region near the sources used in this study. These models are quite similar to one another and both include three crustal layers, in addition to surface sediments. We have taken the model of Mitchell and Hashim (1977) as the crustal compressional wave model for this region and those velocities have been fixed in all of our inversion attempts. Our starting shear wave model for the crust was determined from the compressional wave model by assuming a Poisson ratio value of 0.25. The starting upper mantle model is based upon the model of McEvilly (1964), but has been modified to better explain our group velocity data. The starting shear velocity model appears as a dotted line in each of the models of Figure 6.

Figure 6 presents shear velocity models and their standard deviations resulting from the inversion of the combined Rayleigh- and Love-wave data, from the Rayleigh-wave data only, and from the Love-wave data only. The combined inversion yielded a shear wave value of 1.53 km/sec for the upper 0.25 km of the model and an average value of 3.06 km/sec for the next 0.75 km. Large uncertainties are associated with these values. Resolving kernels for these models are given in Figure 7 and comparisons between the observed and theoretical velocities for the models appear in Figure 8. During the inversion process the

theoretical phase and group velocities were corrected for sphericity according to methods described by Schwab and Knopoff (1972).

As indicated by Figure 6, there are no significant differences between the three models, except perhaps in the lower crust where the shear velocities obtained from the inversion of Love waves (SH) are faster than the shear wave velocities obtained from the inversion of Rayleigh waves (SV) or of the combined data. With the exception of only one value, the upper mantle velocities are always within one standard deviation of the starting model. It is apparent, when the standard deviations and resolving kernels are considered, that an upper mantle low-velocity zone, if present, cannot be resolved by the data. This result was also noted by Biswas and Knopoff (1974). The small velocity decrease at a depth of 65 km occurs because a velocity decrease was present in the starting model at that depth and is not resolvable in the final model.

The relative sharpness of the resolving kernels for crustal depths suggests that we should be able to resolve much more detail in the crust than in the mantle. There are no significant differences among the three models of the crust at depths shallower than about 30 km. At greater depths, however, the shear velocities obtained using Rayleigh waves are lower than those obtained using Love waves. They also exhibit a decrease in value with increasing depth. These results perhaps suggest the possibility of both a low-velocity zone and polarization anisotropy near the base of the crust. Before reaching these conclusions, however, it is necessary to consider the resolving kernels and standard deviations corresponding to that depth range. Figure 7 indicates that the resolving kernels for those depths have a width of about 25 km or slightly more.

Since the thickness of the possible low-velocity layer is less than 20 km, this feature, if present, cannot be resolved using the present fundamental-mode dispersion data.

The theoretical velocities for the three models appear in Figure 8. Almost all of the data agree with the theoretical curves within one standard deviation. The only clear exception is one Love wave phase velocity value at a period of 35 seconds which differs from the theoretical velocity for the model obtained from the combined inversion by a small amount. As might be expected, the theoretical velocity values for the models obtained from separate inversions of the Love- and Rayleigh-wave data correspond more closely to the observed values than do the theoretical values for the model from the combined inversion. Although this better fit could occur because of anisotropy in the crust or upper mantle in the eastern United States, we do not consider that the improvement in fit is substantial enough to warrant that conclusion at the present time.

Synthetic Seismograms

The results of inversion of the fundamental-mode dispersion data
indicate that there are no resolvable differences among the three models
of Figure 6. It might be possible, however, to use higher-mode information
to investigate features of these models im more detail. For example, the
shear velocity at the top of the mantle (S_n) is higher for the model
obtained using both Rayleigh and Love wave (R+L) data than it is for
the model obtained using only Rayleigh wave (R) data. The R+L model
compensates for the higher S_n velocities by having lower shear wave
velocities in the lower crust that those which occur in the R model.

A further question which we might address using the synthetic seismograms
concerns the existence of the low-velocity zone in the lower crust, suggested
by two of the models in Figure 6. As stated earlier, we cannot resolve
this feature using fundamental-mode dispersion data alone. Jordan and
Frazer (1975) have inferred such a low-velocity zone in the lower crust
of eastern Canada from the amplitudes of S to P converted waves at the
base of the crust.

Higher mode data could contribute detailed information on the
shear velocity of the crust if it is of sufficient quality. The higher
mode group velocity data which we obtained, however, exhibits
considerable scatter, and is not of sufficient quality to use for
inversions for structures. We have attempted to study possible detailed
features of our models by computing synthetic seismograms which include
the effects of the fundamental and several higher modes (Tsai and Aki, 1970),
and comparing these to the observed seismograms at several stations. The
synthetics incorporate anelastic attenuation using values from Herrmann
and Mitchell (1975).

Figure 9 shows the observed Rayleigh wave seismograms (vertical component), and the Love wave seismograms (transverse component) obtained from a coordinate transformation of the observed east-west and north-south components, for the October 21, 1965 earthquake, as recorded at four stations. These stations, MDS, AAM, BLA, and ATL, were selected because they are located at azimuths which avoid nodes in the surface wave radiation patterns; thus they provide well-recorded surface waves with a good signal-noise ratio.

The synthetic seismograms for the R+L, the R, and the L models appear in Figure 10. It is apparent that differences between the Rayleigh wave seismograms and between the Love wave seismograms for the different models are not very great. It is also apparent that although the major features of the synthetics agree with those of the observed seismograms, detailed comparisons are not possible. We therefore have no basis for selecting one model over the other, either in comparing model R to R+L, or in comparing model L to R+L. Moreover, there is no reason to propose more than a single isotropic model to satisfy the presently available surface wave data for the eastern United States.

Synthetic seismograms were also computed for models with and without a low-velocity zone in the lower crust. Differences between the synthetic seismograms for the two kinds of models at the various stations were even less substantial than those of Figrue 10. It is not possible therefore, to come to any conclusions concerning the existence or absence of a low-velocity zone in the lower crust of the eastern United States from the presently available surface wave data.

Conclusions

An isotropic model for the crust and upper mantle of the eastern United States can be obtained which satisfies both fundamental-mode Love and Rayleigh wave phase and group velocity data within one standard deviation. Shear velocities in the upper crust of this model increase from 3.4 km/sec to 3.7 km/sec with increasing depth. Shear velocities in the lower crust lie between 4.0 and 3.8 km/sec. Although the model resulting from the inversion process exhibits decreasing velocity with increasing depth in the lower crust, this resulting low-velocity zone cannot be resolved using the data available to us. The uppermost mantle for this model has a shear wave velocity of about 4.8 km/sec. Although a modest low-velocity zone may be present in the upper mantle, one is not required.

Models resulting from separate inversions of the Rayleigh and Love wave data provide slightly better agreement with the observed dispersion data. This result suggests that the lower crust and/or upper mantle of this region could be anisotropic in its elastic properties. The observations are not definitive enough, however, to establish the existence of anisotropy. For the present, therefore, a single isotropic model for the crust and upper mantle of the eastern United States is sufficient to explain all of the surface wave data for that region.

The short-period group velocity data for the eastern United States require the existence of a layer of low-velocity sediments overlying the crust.

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Table 1. Earthquake Parameters (from Herrmann, 1974)

Date	Origin Time	Lat.	Long.	Depth(km)	m_b	T Trend	T Plunge	P Trend	P Plunge
2 Feb. 62	06 43 34.0	36.5	89.6	8	4.3	278	54	52	26
3 Mar 63	17 30 11.4	36.7	90.1	13	4.8	76	31	174	10
14 Aug. 65	13 13 56.6	37.2	89.3	1	3.8	147	1	237	17
*21 Oct. 65	02 04 38.3	37.5	91.0	4	4.9	163	5	305	84
4 June 67	16 14 13.6	33.6	90.9	10	4.5	156	22	248	7
21 July 67	09 14 48.9	37.5	90.4	15	4.3	50	5	314	52
9 Nov. 68	17 01 42.0	38.0	88.5	20	5.5	192	82	97	1
1 Jan. 69	23 35 36.2	34.8	92.6	9	4.4	228	65	330	7
17 Nov. 70	02 13 55.1	35.9N	89.9W	14	4.4	173	32	270	10

*Parameters from this earthquake were taken from Mitchell (1973a).

Table 2

Mean Group Velocities, Phase Velocities, Attenuation Coefficients, and Q values for the Eastern United States

T(sec)	U_R(km/sec)	C_R(km/sec)	$*\gamma_R$(km^{-1})	Q_R	U_L(km/sec)	C_L(km/sec)	$*\gamma_L$(km^{-1})	Q_L
4	2.894 ± .201				3.258 ± .087			
5	2.950 .102		5.10 x 10^{-4}	418	3.348 .087		7.20 x 10^{-4}	261
6	2.996 .073		5.50	318	3.399 .077			
7	3.036 .070		5.87	252	3.444 .084			
8	3.075 .064		4.06	315	3.434 .071		3.35	341
9	3.087 .057		2.45	462	3.440 .070		2.07	490
10	3.104 .059	3.301 ± .045	3.12	324	3.443 .070	3.629	0.44	2074
12	3.099 .058	3.373 .034	2.70	313	3.457 .062	3.712 ± .039	-0.05	
14	3.092 .061	3.440 .030	1.37	530	3.492 .035		0.07	9180
16	3.088 .067	3.490 .039	1.76	361	3.453 .055	3.807 .012	0.80	711
18	3.097 .068	3.529 .046	1.01	558	3.476 .058	3.914 .035	0.67	749
20	3.135 .076	3.582 .052	1.19	421	3.474 .049	3.990 .018	1.01	448
22	3.140 .077	3.610 .060	1.47	309	3.491 .045	3.984 .075	1.02	401
24	3.177 .069	3.657 .061	1.50	275	3.512 .041	4.068 .048	0.84	444
26	3.225 .094	3.725 .072	1.06	354	3.534 .049	4.137 .083	0.59	580
28	3.251 .084	3.786 .070	1.27	272	3.534 .049	4.163 .065	0.60	529
30	3.362 .098	3.811 .079	1.31	238	3.549 .050	4.141 .065	0.93	317
35	3.515 .127	3.916 .113	0.75	341	3.584 .067	4.350 .071	0.46	544
40	3.580 .143	4.021	1.25	176	3.653 .067	4.421 .084	1.13	190
45	3.745 .136	3.913			3.748 .073	4.434	0.02	
50	3.856 .072	4.071			3.865 .065	4.562		9313
60	3.870				3.960 .132			
70	3.864							

*From Herrmann and Mitchell (1975)

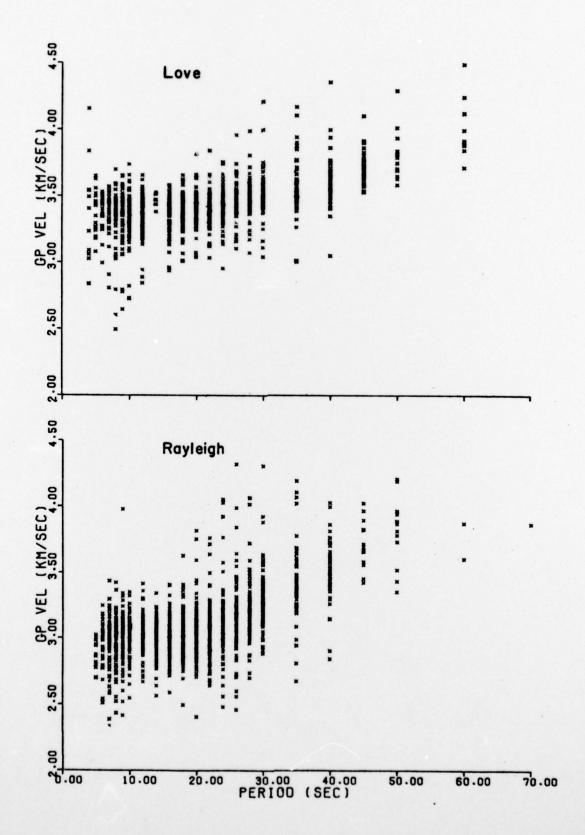Standard deviations for U and C are not given for cases with two or fewer observations.
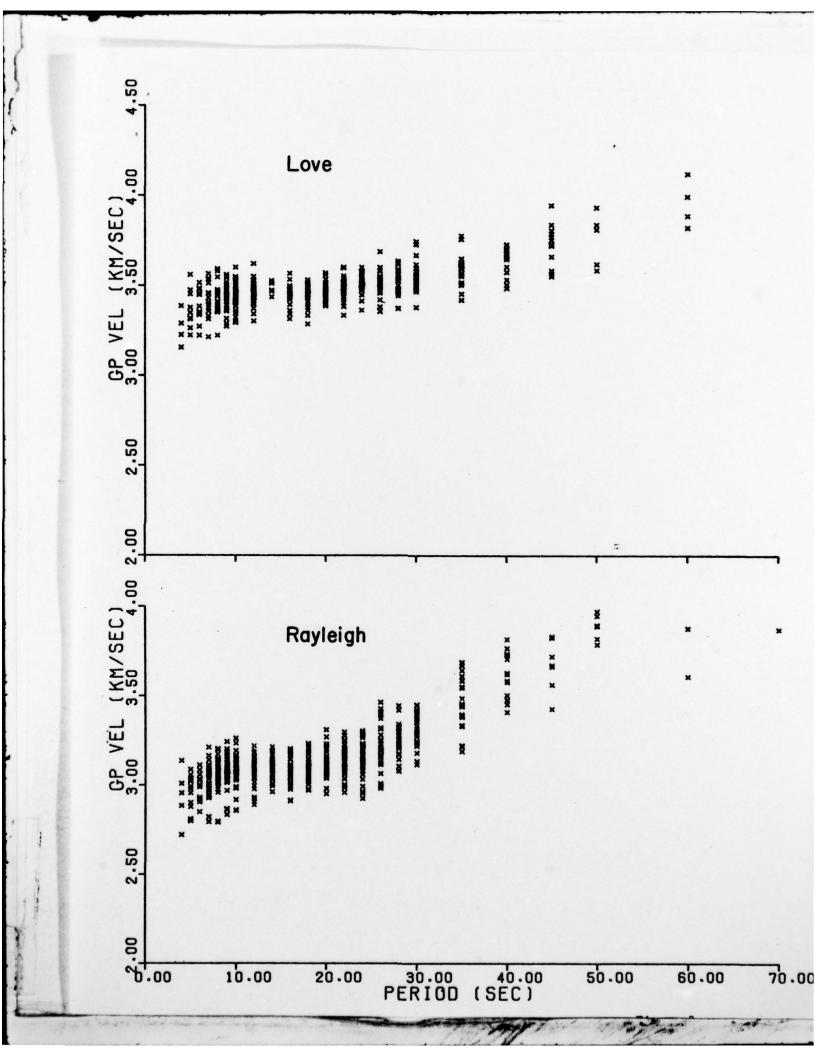
Figure Captions

Figure 1. Map indicating the location of earthquake sources and seismograph
stations in eastern North America from which data used for surface wave
inversion were obtained.

Figure 2. All group velocity data obtained using the central United
States earthquakes of Table 1. These data were obtained from
numerous stations throughout eastern and western North America
without regard to regionality or signal-noise ratio.

Figure 3. Group velocity data which pertains only to eastern North
America between the sources and seismograph stations shown in
Figure 1. These data remain after windowing to exclude paths
outside the distance range 300-2000 km, and outside the azimuth
range 340° - 135°. Paths along azimuths from nodes in the
surface wave radiation patterns have also been excluded.

Figure 4. Theoretical amplitude spectra for the fundamental-mode (F)
and nine higher-mode Love waves generated by the earthquake of
October 21, 1965 at a depth of 4 km (left), and by an earthquake
with the same fault-plane solution, but at a depth of 15 km (right).

Figure 5. Phase and group velocity data and the standard deviations which
were used for inversion plotted with theoretical values for
models of McEvilly (1965). The solid lines, the short dashes, and
the long dashes pertain to models which are isotropic, anisotropic,
anisotropic in the upper mantle, and anisotropic in the lower
crust and upper mantle, respectively.

Figure 6. Shear velocity models with standard deviations which have resulted
from the inversion of both Rayleigh and Love waves (left), Rayleigh waves
only (center), and Love waves only (right). The dotted lines denote the
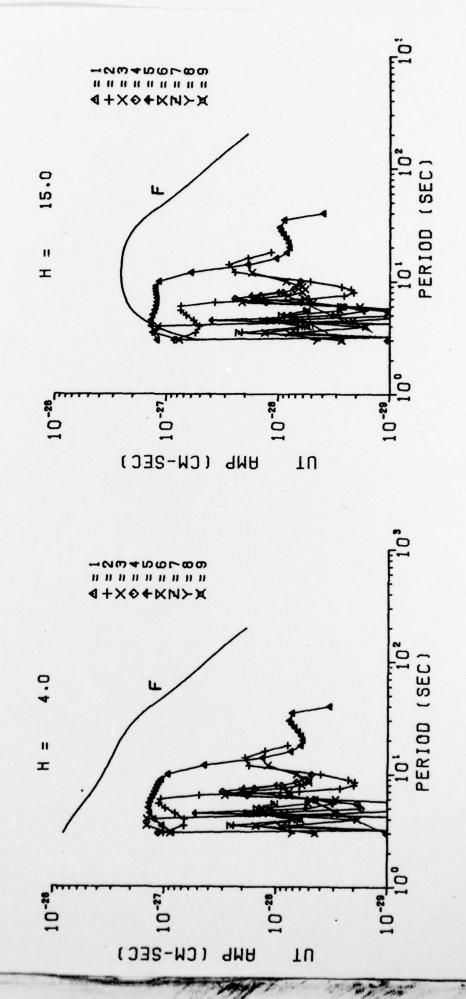starting model used for the inversion process.

Figure 7. Resolving kernels for the models of Figure 6.

Figure 8. Phase and group velocity data used for inversion and
theoretical values pertaining to the models of Figure 6. The
solid lines correspond to the inversion of the combined Love-
and Rayleigh-wave data and the dashed lines correspond to
separate inversions of the Love- and Rayleigh-wave data.

Figure 9. Seismograms recorded at stations MDS, AAM, BLA, and ATL for
the earthquake of October 21, 1965. The seismograms for transverse
motion were obtained from a coordinate transformation of the
east-west and north-south components. The zero times correspond
to the beginning of digitization.

Figure 10. Synthetic seismograms corresponding to the event and
stations of Figure 9 for models obtained from a simultaneous
inversion of Rayleigh and Love wave data (left), from an inversion
of Rayleigh data only (upper right), and from an inversion of Love
wave data only (lower right). The starting times for each three-
minute segment do not necessarily correspond to the starting times
in Figure 9. The two second noise results because a boxcar
filter with a sharp cutoff at 0.5 hz was used in the generation of
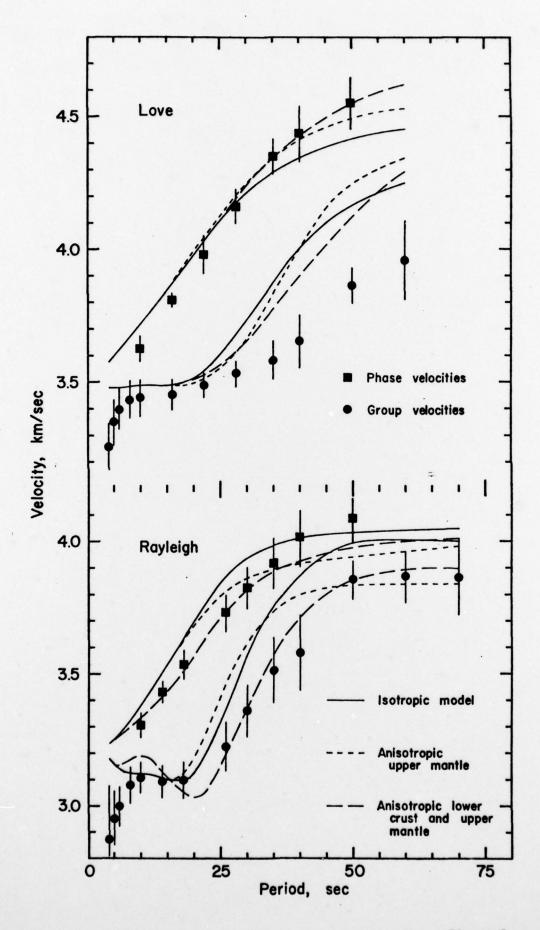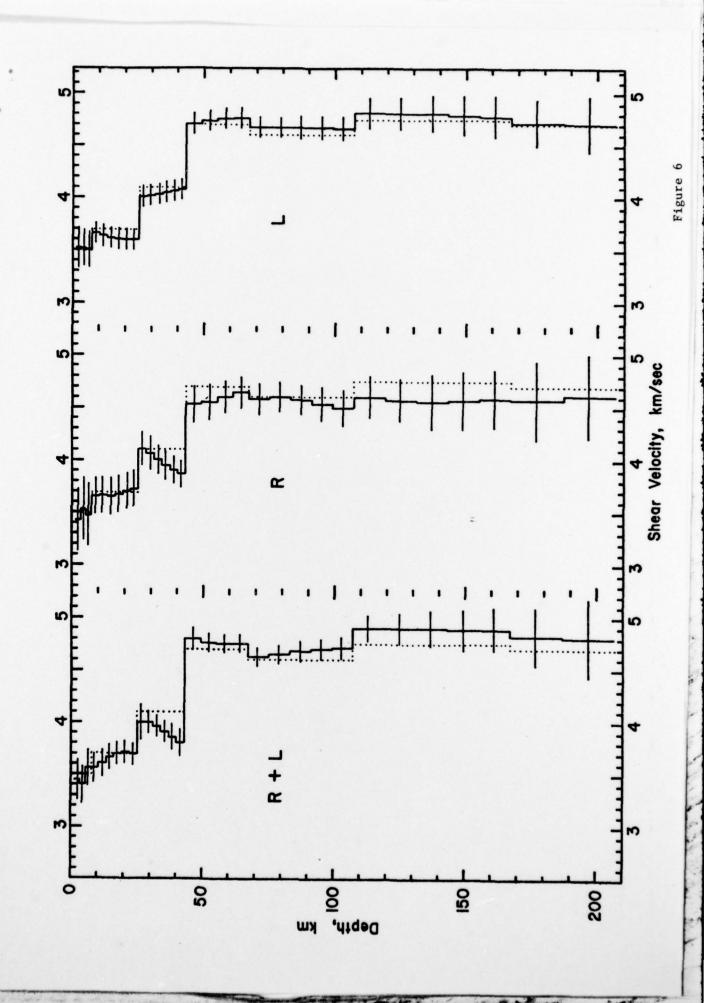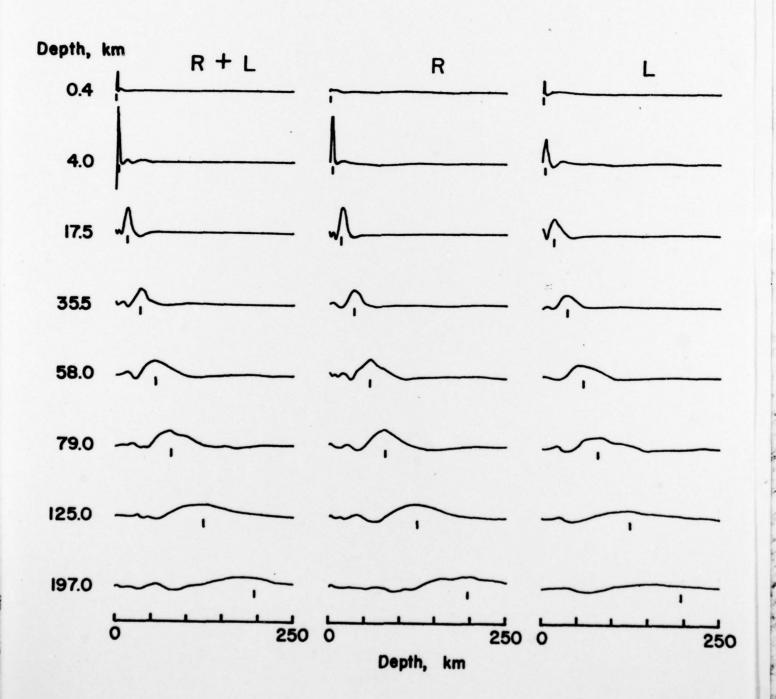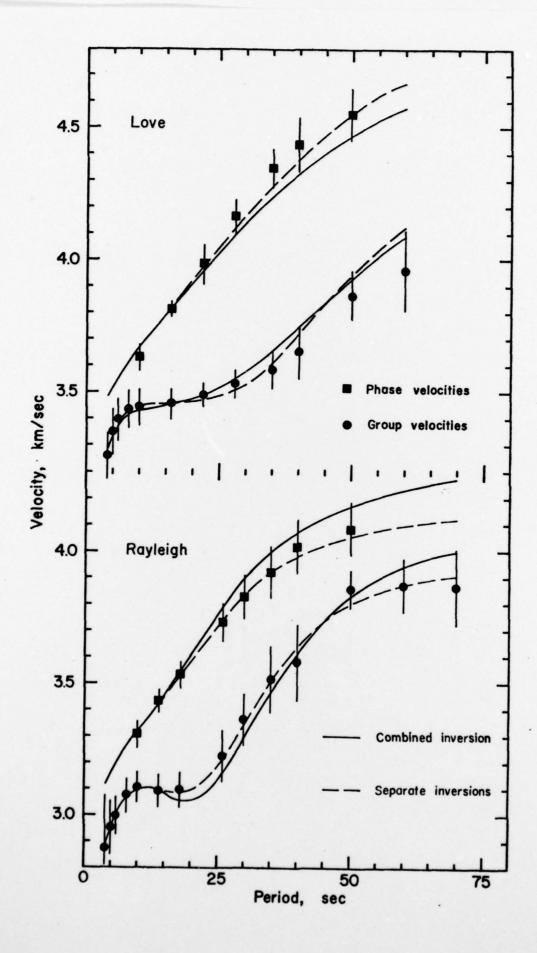the seismograms.

Figure 1

Figure 2

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10